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Soil hydromorphy and soil carbon: A global data analysis

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ABSTRACT

Wetland soils are an important component of the Global Carbon Cycle because they store about 20–25% of the terrestrial soil organic carbon (SOC). Wetlands occupy about 6% of the global land surface and any change in their use or management has potentially dramatic consequences on greenhouse gases emissions. However, the capacity of wetland soils to store carbon (C) differs from place to place due to reasons still not well understood. The objective of this review was to evaluate the global variations in wetlands SOC content (SOC_C) and to relate it to key soil and environmental factors such as soil texture, intensity of soil hydromorphy, metallic element content and climate. A comprehensive data analysis was performed using 122 soil profiles from 29 studies performed under temperate, humid, sub-humid, tropical and sub-arctic conditions. The results point to average SOC_C of 53.5 ± 15.8 g C kg^{-1} with a maximum of 540 g C kg^{-1}. SOC_C increased with increase in intensity of soil hydromorphy ($r = -0.52$), Al ($r = 0.19$) and Fe content ($r = 0.21$), and decreased with soil pH ($r = -0.24$). There was also a surprising tendency for intensity of soil hydromorphy, and thus SOC_C, to decrease with increasing mean annual precipitation and soil clay content. These results contribute to a better understanding of the impact of soil hydromorphy in wetlands on organic C stabilization in the soils. However, further studies with additional information on soil bulk density to assess carbon C stocks, still need to be performed.

1. Introduction

Wetlands are transitions from aquatic to dry land (upland) habitats and they occur in areas where soils are inundated/saturated by water due to high groundwater or ponding surface water during a part or all of the year (Neue et al., 1997). Globally, wetlands cover over 2.1 million km^2 (Ramsar Convention, 2016), representing about 6% of the earth's surface (Zhang et al., 2016). Even though wetlands provide a wide range of ecosystem services of great social and economic values, their rate of deterioration is faster than that of any other ecosystem (Davidson and Finlayson, 2007). For example, US wetland area was reported to have declined by about 25,200 ha between 2004 and 2009 (US Fish and Wildlife Services, 2011). Wetlands are also a highly threatened ecosystem type in New Zealand with only < 10% of their original extent remaining nowadays (Myers et al., 2013).

All wetland soils are characterized by excess water on momentary to permanent bases (EPA, 2017). The excess water induces reduction processes (Vepraskas and Guertal, 1992) resulting in generally low soil chroma and value (< 2). Additionally, the lack of oxygen lessens organic matter mineralization with subsequent accumulation of organic carbon (C) in the soils (Vaughan et al., 2008). Gorham (1995) and Zhang et al. (2016) estimated worldwide wetlands' C storage to range from 150 to 535 Gt C, which makes wetlands a key component of the global C cycle. The capability of wetlands to accumulate organic C varies with climate, soil properties and hydrological regime (Zhang et al., 2011; Wang et al., 2016). Accumulation of C in wetlands is a balance between net primary productivity and decomposition mediated by microbial processes (Stemmler and Berthelin, 2003; Zerva and Mencuccuni, 2005), decomposer community, environmental conditions and substrate litter quality (Werker et al., 2002; Koide et al., 2005; Urakawa and Bernhard, 2017).

Wetlands of cold regions have the greatest C accumulation potential due to low temperatures, high humidity and weak microbial activities (Tockner and Stanford, 2002) with, for example, boreal peatlands containing 270–370 Gt C (Fan et al., 2013), compared to tropical wetlands which stock only 89 Gt C (Page et al., 2011) because of higher temperatures and microbial activities. The boreal and temperate wetlands are C sinks (Moore and Bellamy, 1974) and they accumulate peat layers at rates regulated by climate, vegetation and topography (Armentano and Menges, 1986). The total organic C pool in the

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permafrost zone is composed of C frozen at depth in peatlands and C intermixed with mineral soils (Schuur et al., 2008). The northern permafrost zone alone is estimated to store 1672 Pg C; 277 Pg C of that in peatlands and 1024 Pg C in the 0–3 m permafrost-zone, representing a large fraction of global soil C stocks (Schuur et al., 2008). Thawing of the permafrost, which results in decomposition of previously frozen organic C, constitutes a significant feedback from terrestrial ecosystems to the atmosphere (Schuur et al., 2008) with approximately 100 Pg C set to be released from thawing permafrost by 2100 if temperature increases by 4 °C (Gruber et al., 2004). The collapsing permafrost generates fens in hydromorphic zones and bogs in the drier zones. Trumbore et al. (1999) showed that different decomposition rates give rise to the greatest C accumulation in the intermediate zone. Accumulation of C in the bogs and fens is low due to low inputs matched by low outputs and high productivity matched by high respiration, respectively. The collapse of permafrost wetlands also changes plant and microorganism communities with, for example, increasing organic C degraders in Alaska (Chapman et al., 2017).

On the other hand, tropical wetlands are the largest natural source of atmospheric C with organic substrate quality the main control of decomposition (Miyajima et al., 1997). Substrate quality is important in both anaerobic decomposition and methanogenesis, and correlate strongly with lignin content (Miyajima et al., 1997; Hoyos-Santillan et al., 2016). Hoyos-Santillan et al. (2016) reported high CO_2 and CH_4 emissions under anoxic conditions in tropical peatlands, which strongly correlated to lignin content, long chain fatty acids and polysaccharides concentrations. Unlike temperate and boreal peatlands which come from more degradable sphagnum mosses and sedges (Yavitt and Williams, 2015a), tropical peatlands come from high lignin tree litter. Although decomposition rates are generally faster in lower C:N ratio substrates, C:N ratio had no limit on mineralization in tropical wetlands (Yavitt and Williams, 2015b). However, Sjogersten et al. (2016) showed no link between organic chemistry of arctic and subarctic peatlands and CH_4 emission under flooded and non-flooded conditions. The emission was limited by availability of specific substrates i.e. sugars and low molecular weight organic acids, which ferment to acetate, a precursor for CH_4 production. Unlike high latitude wetlands with relatively stable water levels (Junk et al., 2014), neotropic peatland C losses follow water level drawdown and litter quality (Bridgham and Richardson, 1992; Laiho, 2006). The neotropic wetland hydrology, characterized by river flooding and groundwater table fluctuations (Junk et al., 1989), modifies the chemical and physical properties, degree of anoxia, sediment properties and soil pH, which subsequently reflect in the decomposition rates of organic materials (Webster and Benfield, 1986).

However, there are discrepancies in wetland C storage for any given region because decomposition in the wetlands is a complex aerobic-anaerobic process regulated by physical, chemical, and biological processes (Kayranli et al., 2010). Landscape diversity also contrasts wetlands by influencing vegetation growth and C storage capacity (Heinselman, 1970). In similar Brazilian climatic conditions, for example, Martins (2014) pointed to maximum SOC content (SOC_C) of 41 g C kg^{-1}, while Bardy et al. (2008) and Rosolen et al. (2015) estimated maximum values of 455 and 537 g C kg^{-1}, respectively. Chaplot et al. (2001) and Rosolen et al. (2015), amongst others, also reported on large differences in SOC_C of a single wetland, which may be associated with differences in soil saturation by water and hydromorphy. The results of Chaplot et al. (2001) showed the highest SOC_C of 460 g C kg^{-1} at the valley bottom and a lowest of 100 g C kg^{-1} at the footslope in the Armorican Massif of western France.

Relationships between flooding duration and frequency and decomposition rates in wetlands are still not well understood due to great variability in wetland types (Brinson et al., 1981; Neiff et al., 2006). The present study aimed at (i) reviewing existing works that assessed SOC_C levels in wetland soils at a global scale and (ii) investigating the links between SOC_C and soil hydromorphy and the main soil and environmental factors of control. These objectives were achieved by gathering quantitative data from various natural wetlands in different environments worldwide. Such quantitative analysis is important for land managers concerned with understanding the potential impact of change in wetland hydromorphy through, for instance drainage, on C losses from the soils. It might also help in estimating the C sequestration potential of degraded wetland soils in the framework of international programs such as the '4 per mille Soils for Food Security and Climate', which was launched at the COP21 with an aspiration to increase global soil organic matter stocks. The results could in particular inform on ways to render certain wetlands more efficient in sequestrating atmospheric C. Finally, quantitative information on the links between, for instance, climate or soil clay content and wetland SOC_C could constitute an important tool for the spatial prediction of wetland soil C.

2. Materials and methods

2.1. Study setup

The study was based on data collected from electronic reserves and databases containing field experiment reports performed on wetlands or waterlogged sites. The literature data search targeted information from soil profiles developed under different wetland types e.g. floodplains, swamps and peatlands. The search used key words and phrases such as "water-logged soil", "soil hydromorphy", "hydromorphic features", "*Gleysols*", and "SOC". A number of electronic databases such as Google, Google Scholar, Web of Science and Science Direct were searched for the data. The methods used in organizing the database and statistical treatments of the data were according to Mutema et al. (2015).

Many papers were downloaded but gaps in data on SOC_C and/or information on the intensity of soil hydromorphy led to omission of some papers from the current analysis. Other important soil parameters sought included iron (Fe), aluminum (Al), clay content and Munsell soil color (Hue, Value and Chroma). In addition to these parameters, the data sources also needed to provide site locational and climatic information. Boreal and temperate wetlands were not included the current study because they accumulate very large amounts of C and would behave as outliers when put together with other environments. Outliers are values that shift the median and mean up. The final database consisted of 27 published ISI papers and two masters' theses, totalizing 122 soil profiles from different sites across the world. Amongst these accepted data sources, 12 came from Brazil, 3 from Germany, two from China and USA, and one from each of Switzerland, Bangladesh, Italy, India, Mexico, Bolivia, Russia, South Africa, and France (Fig. 1) with sites spanning a latitudinal range from 30° 15′ S to 47° 57′ N.

The database captured information on author name(s), years the papers were published, paper titles, experimental site characteristics such as geographical location, climatic parameters, dominant land use, number and descriptions of the studied soil profiles. The data on SOC_C, Clay, Fe and Al, collected from the data sources, were in different units but mainly presented in terms of mass (e.g. g kg^{-1}, mg g^{-1} and percentage %). The data were normalized to g C kg^{-1} for SOC_C and % for Fe, Al and Clay. Other site characteristics recorded include longitude (LONG), latitude (LAT) in degrees, 30-year mean annual precipitation (MAP) and temperature (MAT). Table 1 presents a summarized version of the database.

2.2. Definitions of variables

2.2.1. Climatic factors

The climatic factors considered in the present study were MAP and MAT. They were grouped into categories (Table 2) following Mutema et al. (2015). MAP was categorized into 120–800, 800–1200 and 1200–2200 mm $year^{-1}$, while MAT was grouped into 6–15, 15–22 and 22–30 °C $year^{-1}$.

2.2 Soil properties.

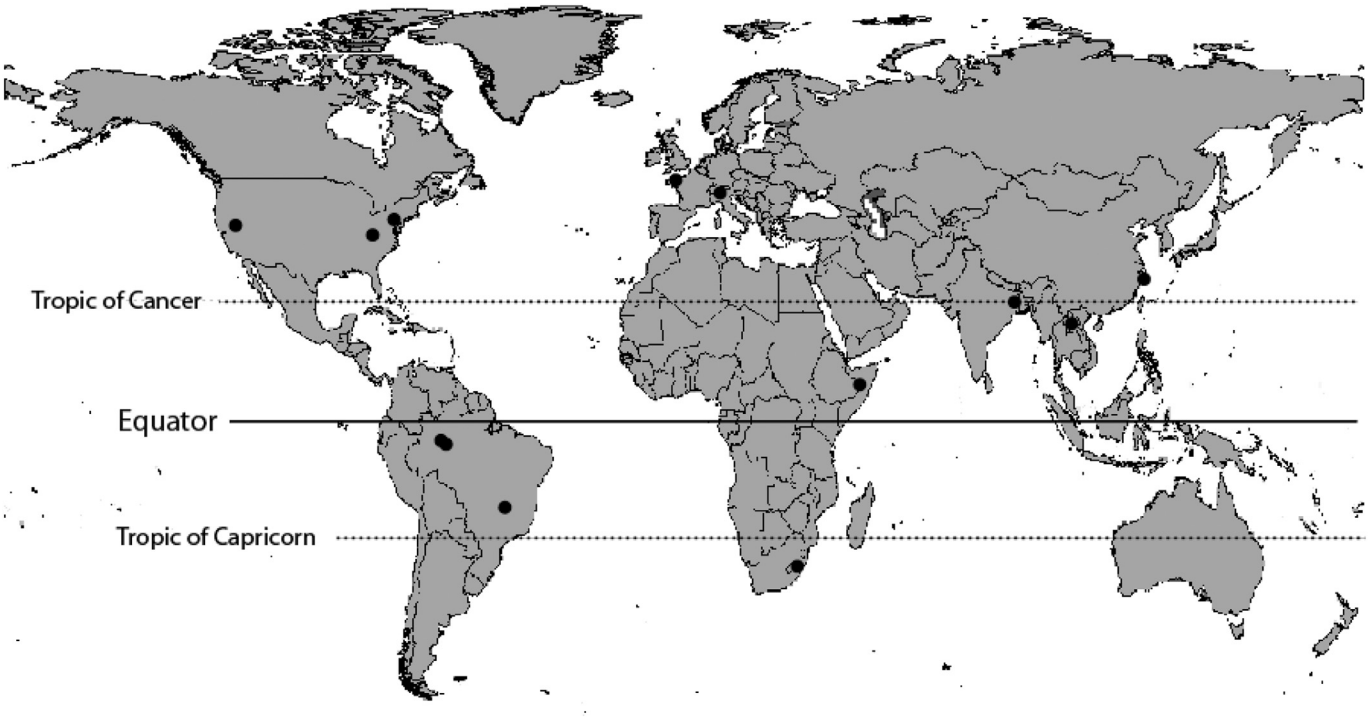

Fig. 1. Global distribution of the study sites used in this meta-analysis.

Soil properties, such as soil clay content, intensity of soil hydromorphy, soil Munsell Value and Chroma, soil pH and soil depth also varied greatly from one site to another and were grouped into three categories each (Table 2). Soil clay content categories were adopted from Mutema et al. (2015), namely 0–25%, 25–50% and > 50%. Chroma was grouped into 0–4, 4–7 and 7–10, and value into 2–4, 4–6 and 6–8. The Hydromorphic Index HI (Chaplot et al., 2001) was grouped into three categories, namely 0–15 for low intensity of soil

Table 1
Compilation of references included in the database for analysis of the factors controlling SOC in hydromorphic soils.

Author(s)	Country	LONG	LAT	MAP	MAT	SOC_C			Clay	Chroma	Value
		°		mm year^{-1}	°C year^{-1}	g C kg^{-1}			%		
						Mean	Min	Max			
Bardy et al. (2008)	Brazil	−1.87	−61.60	2000.00	26.00	319.80	165.00	455.00		1.57	3.32
Brinkman (1977)	Bangladesh	90.09	24.50	2923.00	23.90	2.63	0.20	6.50	7.20	1	6.44
Bullinger-Weber et al. (2014)	Swiss	8.60	47.56	870.00	10.10	20.19	1.10	56.20	0.70	–	–
Chaplot et al. (2001)	France	5.3	47.5	1000.00	10.8	56.86	1.2	460	1.87	2.77	4.17
Ferronato et al. (2016)	Italy	12.24	44.51	600.00	13.50	18.61	1.10	118.10		1.13	3.75
Fiedler et al. (2002)	Germany	8.12	47.12	785.00	8.30	209.06	2.00	540.00	18.84	3.08	4.45
Fiedler and Kalbitz (2003)	Germany	47.12	8.12	1600.00	6.00	157.57	5.00	430.00	11.40	–	–
Fritsch et al. (2009)	Brazil	−1.87	−61.60	2000.00	26.00	3.20	1.65	4.55		–	–
Gangopadhyay et al. (2015)	India	23.02	91.58	2150.00	18.70	8.35	0.50	24.30	38.90	1.71	4.88
Guimaraes et al. (2013)	Brazil	−3.17	−60.50	2000.00	25.00	7.36	0.18	36.27	21.40	7.36	5.07
Jungkunst et al. (2008)	Germany	47.95	8.12	1600.00	6.00	274.53	23.00	500.00	25.74	–	–
Kölb et al. (2014)	China	30.17	121.23	1325.00	13.60	7.93	1.40	30.00	28.94	2.53	4.02
Machado (2014)	Brazil	−19.36	−48.00	1474.00	22.55	89.00	17.77	175.88	57.36	–	–
Martins et al. (2006)	Brazil	−49.50	−10.66	1750.00	24.00	29.13	0.30	121.10	29.88	3.17	5.52
Martins (2014)	Brazil	−19.22	−48.06	1474.00	22.55	13.44	1.42	41.27	56.96	2.09	5.3
Mello et al. (2015)	Brazil	−16.50	−56.17	1150.00	25.00	13.51	6.40	23.83		–	–
Nascimento et al. (2004)	Brazil	−1.87	−61.60	2000.00	26.00	19.35	0.60	54.30	8.48	2.33	6
Nascimento et al. (2010)	Brazil	−16.80	−48.68	1440.00	23.00	12.64	0.69	40.86	2.50	–	–
Neto et al. (2015)	Brazil	−29.75	−50.50	1309.00	19.50	46.77	1.90	453.00	18.45	2.38	3.76
Olaleye et al. (2014)	South Africa	−28.88	−28.78	464.00	15.50	6.35	3.63	8.89	3.83	1.62	2.91
Orozco et al. (2014)	Mexico	19.64	−101.59	919.00	16.80	26.14	4.00	66.00	26.21	2.42	2.69
Rosolen et al. (2015)	Brazil	−18.93	−47.83	1516.00	23.00	31.62	5.00	116.00	537.6	2.59	3.88
Schawe et al. (2007)	Bolivia	−16.15	−68.92	3833.30	13.30	120.36	11.00	447.00	14.64	–	–
Schiavo et al. (2012)	Brazil	−18.99	−56.60	1433.00	30.15	7.95	0.10	110.50	7.47	3.52	4.52
Skorupa et al. (2013)	Brazil	−48.02	−15.75	1594.00	21.20	41.11	3.00	190.00	34.42	3.06	3.44
Sorokin et al. (2006)	Russia	106.36	57.28	244.00	−1.60	19.59	2.50	41.20	9.11	–	–
Thompson et al. (2007)	USA	−83.40	36.80	127.00	12.50	27.18	2.0	65.00		1.14	4.37
Vepraskas and Caldwell (2008)	USA	35.57	−77.44	1260.30	16.40	22.06	2	71		2.11	4.17
Wang et al. (2011)	China	37.58	−118.55	552.00	12.10	57.93	35.40	107.50		–	–

Table 2
Environmental factors, soil variables and their classes as used in the current paper.

Environmental factor	Symbol	Remarks	Class definition	Class name
Mean annual precipitation (mm year^{-1})	MAP		120–800	Semi-arid
			800–1200	Humid
			1200–2200	Wet
Mean annual temperature (°C)	MAT		6–15	Cold
			15–22	Warm
			22–30	Hot
Soil clay content (%)	Clay		10–25	Sandy loam
			25–50	Sandy Clay
			50 >	Clay
Chroma	Chroma		0–4	Low
			4–7	Medium
			7–10	High
Value	Value		2–4	Low
			4–6	Medium
			6–8	High
Hydromorphic index			0–15	Low
			15–20	Medium
			20–50	High

hydromorphy, 15–20 for moderate intensity and 20–50 for high intensity. Low soil hydromorphy corresponded to soils with episodic to no saturation by water, moderate intensity to soils periodically saturated, and high intensity to soils temporally to permanently saturated by water (Chaplot et al., 2001; Franzmeier et al., 1983). Soil depth was also grouped into three domains, namely 0–40, 40–80 and 80–120 cm. The SOC, Al and Fe contents used in the current analysis referred to average values obtained from amounts present in horizons A and B of the soil profiles.

2.3. Data analyses

The sample size (N) for each environmental factor and response variable was first determined, and then summary statistics, described by minimum, maximum, median, standard deviation (SD), skewness, 25th quartile (Q1) and 75th quartile (Q3), kurtosis and coefficient of variation (CV%), generated (Table 3). Variability and distribution of the datasets for the different factor categories were elucidated using box-plots (Figs. 2–3). Each box-plot shows the minimum, maximum, median, Q1 and Q3 values after outliers were checked and removed from the box-plots. Spearman correlation analyses (Table 4) were performed to evaluate the one-on-one relationships between controlling factors and response variables. Finally, principal component analysis was used to depict the multiple relationships between the controlling environmental factors and response soil variables (Fig. 4). A principal component analysis (PCA) is a powerful tool that visually presents linear relationships between variables by converting non-linear

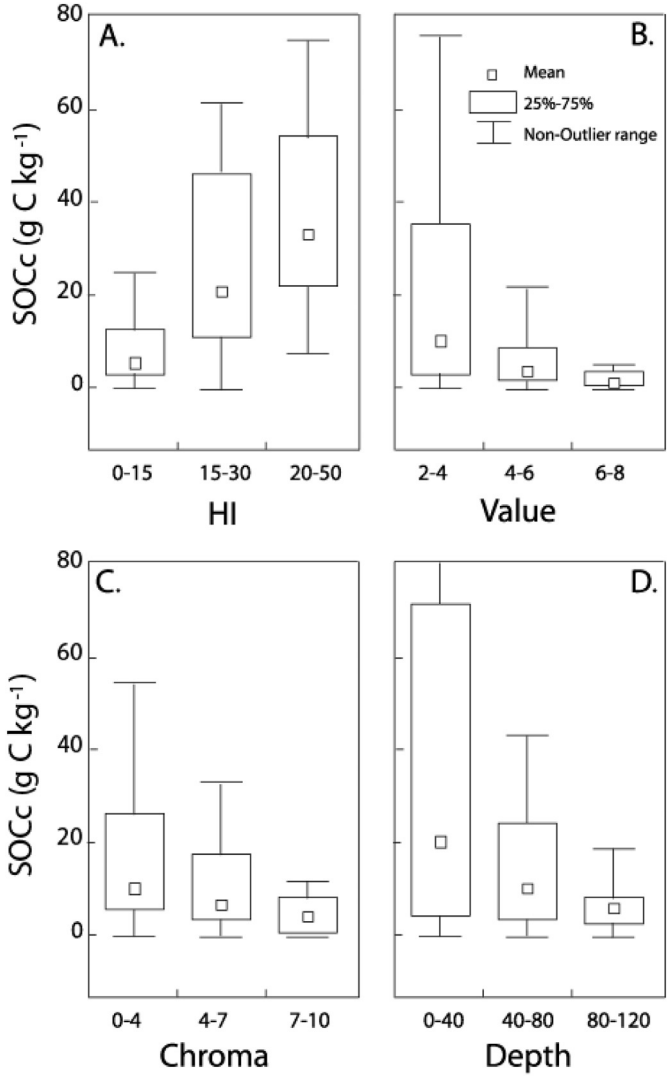

Fig. 2. SOCc as function of (A) soil hydromorphy (HI index), (B) soil Value, (C) Chroma and (D) Depth.

Table 3
Summary statistics of the environmental and soil variables.

Statistic	Controlling factors						Soil variables		
	MAP	MAT	pH	Clay %	Chroma	Value	SOC_C	Fe	Al
N	28	28	453	502	505	505	610	126	115
Min	127.00	−1.60	2.50	0.20	0	1.00	0.01	0	0
Max	3833.3	30.15	8.90	83.61	10.00	8.00	54.00	17.3	38.3
Mean	1316.6	16.74	5.71	21.34	2.53	4.27	5.35	2.10	3.67
SEM	24.24	0.26	0.07	0.83	0.08	0.06	0.44	0.27	0.80
Median	1325	15.5	5.25	19.00	2.00	4.00	1.18	1.17	0.99
Quartile 1	785.00	12.5	4.67	5.03	1.00	3.00	0.42	0.50	0.64
Quartile 3	1600	23.00	7.20	29.20	3.00	5.00	3.93	1.90	1.61
Stdev	648.28	6.70	1.55	18.28	1.87	1.40	10.89	3.08	8.66
Skewness	1.01	−0.02	0.15	0.99	1.54	0.28	2.92	2.89	3.27
Kurtosis	2.97	−0.59	−0.89	0.47	2.03	−0.65	7.65	9.40	9.40
CV%	49.23	40.04	27.13	85.69	73.88	32.78	203.52	146.91	235.72

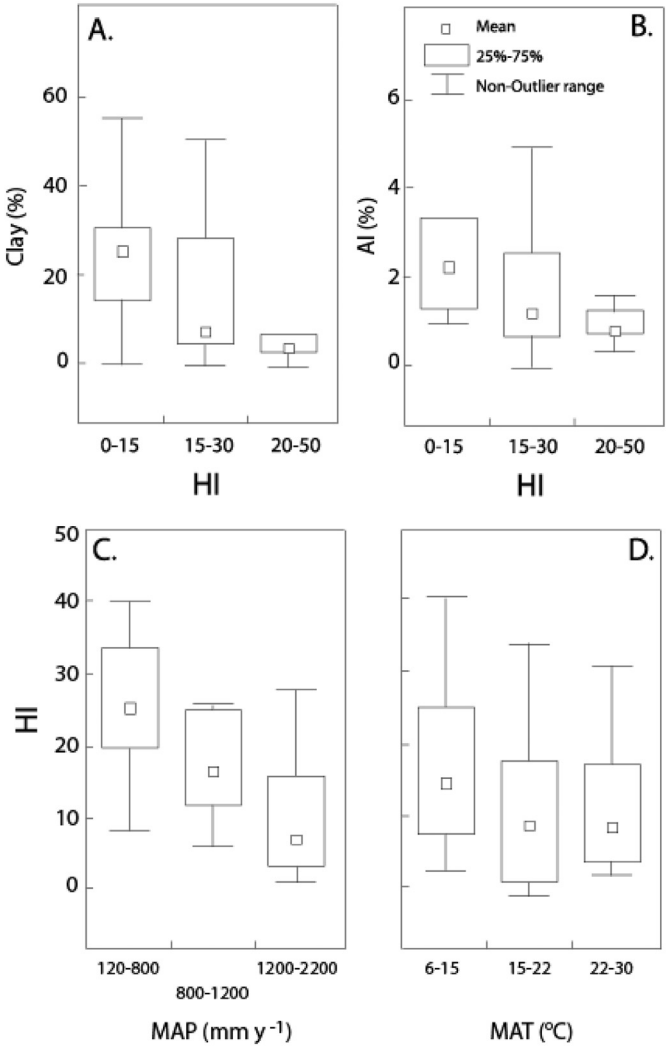

Fig. 3. (A) Clay and (B) Al as function of HI, and HI as a function of (C) MAP and (D) MAT index.

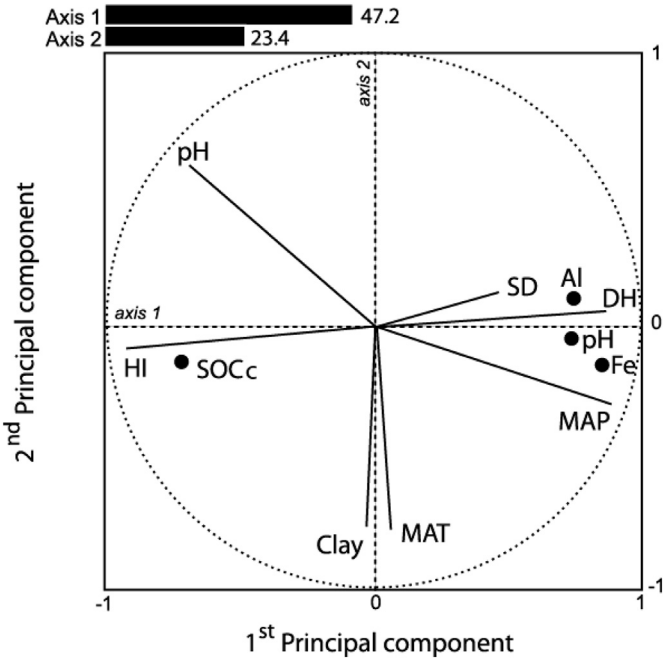

Fig. 4. Principal Component Analyses (PCA) of environmental factors (MAP: mean annual precipitation; MAT: mean annual temperature; SD: soil depth; DH: depth of hydromorphy in the soil profile) and soil parameters (Clay: soil clay content, SOC_C: soil organic carbon content, Chroma and Value, pH and, Al and Fe content).

variables into related linear components (Jambu, 1991).

3. Results

3.1. Variability of environmental factors and soil variables

The summary statistics in Table 3 indicated that the environmental and biogeochemical parameters from the wetland soil profiles varied widely across the different locations. MAP ranged from minimum 127 mm $year^{-1}$ in USA (Thompson et al., 2007) to maximum 3833 mm $year^{-1}$ in a Bolivian tropical forest (Schawe et al., 2007), with an average of 1316 mm $year^{-1}$. MAT varied from minimum −1.6 °C $year^{-1}$ in temperate Russia (Sorokin et al., 2006) to 30.15 °C in tropical Brazil (Schiavo et al., 2012). Soil pH was very acidic with a value of 2.5 in an alluvial wetland in Italy (Ferronato et al., 2016), while the other extreme was the basic soil with a pH value of 8.9 in a floodplain of a tropical forest in Brazil (Guimaraes et al., 2013). Soil clay content varied from a minimum of 0.20% at a site in Brazil (Martins, 2014) to a maximum of 83% in Bangladesh (Brinkman, 1977). The overall average soil clay content was 21%. Soil color also showed high variability with a respective coefficient of variation of 33% for Value and 75% for Chroma. Chroma ranged from 0 for some sites in Brazil, USA and China (Skorupa et al., 2013; Vepraskas and Caldwell, 2008; Kölb et al., 2014) to a maximum of 10 for a site in Brazil (Schiavo et al., 2012). Value varied from a minimum of 2 at sites in the US, Germany, Brazil and China (Vepraskas and Caldwell, 2008; Fiedler et al., 2002; Nascimento et al., 2004; Kölb et al., 2014) to a maximum of 8 reported for a site in Brazil (Martins et al., 2006). Soil variables varied more widely than environmental controlling factors with coefficients of variation of 265, 147 and 236% for SOC_C, Fe and Al, respectively.

Table 4
Spearman rank correlations between environmental factors and soil variables.

	LONG	LAT	MAP	MAT	pH	Clay	Chroma	Value	SOC_C	Fe	Al
LONG	1.00										
LAT	0.36*	1.00									
MAP	−0.17*	−0.31*	1.00								
MAT	−0.47*	−0.74*	0.51	1.00							
pH	−0.09*	−0.14*	−0.09	0.05	1.00						
Clay	−0.01	−0.08	0.03	0.09	0.21*	1.00					
Chroma	0.09*	−0.05	0.06	0.03	0.18*	0.19*	1.00				
Value	−0.01	−0.08	0.04	0.06	0.23*	−0.03	0.39*	1.00			
SOC_C	0.034	0.15*	−0.02	−0.14*	−0.24*	−0.36*	−0.26*	−0.42*	1.00		
Fe	−0.00	0.13	−0.12	−0.35*	−0.11	−0.04	0.17	−0.01	0.21*	1.00	
Al	0.09	−0.10	−0.09	−0.06	0.05	0.19	0.67*	0.22	0.19*	0.62*	1.00

* Significant correlation at $p < 0.05$.

Minimum SOC_C was 0.1 g C kg^{-1} in South Africa (Olaleye et al., 2014) and a maximum of 540 g C kg^{-1} was observed in Germany's Black Forest, located in humid temperate climate (Fiedler et al., 2002). The lowest Fe and Al contents of close to 0% were found in Russian (Sorokin et al., 2006). The highest soil Fe content was 17% found in Italy (Ferronato et al., 2016), while the highest Al content of 38% was reported in Brazil by Fritsch et al. (2009).

3.2. Variation of SOC_C as a function of soil and environmental factors

Box-plots in Fig. 2 show the variability of SOC_C as a function of hydromorphic index (HI), soil Value, soil Chroma and soil depth. SOC_C showed a tendency to increase with HI (Fig. 2A). On the basis of median values, SOC_C increased from 5.5 to 20 g C kg^{-1} (an increase of 263%) as HI increased from 0 to 15 to 15–30 (Fig. 2A). This was followed by a further 55% increase to 31 g C kg^{-1} for the 20–50 HI range. In contrast, SOC_C tended to decrease with soil Value, soil Chroma and soil depth. The decrease from Value range 2–4 to 4–6 was from 9.5 to 3.1 g C kg^{-1} and this was followed by a further decrease of 2 g C kg^{-1} from 4 to 6 to 6–8 (Fig. 2B). SOC_C also showed a general decrease with increase in soil Chroma (Fig. 2C). It decreased from 9.1 to 6.0 g C kg^{-1} from 0 to 4 to 4–7 Chroma, and to 3.5 g C kg^{-1} for 7–10 Chroma. The decrease of SOC_C from the top (0–40 cm) to subsurface (40–80 cm) layer was greater than from the subsurface to deeper (80–120 cm) soil layer (Fig. 2D). The SOC_C decreased by 48% from the top to subsurface layer (i.e. from SOC_C = 20 to 9.2 g C kg^{-1}), followed by 45.6% decrease from the subsurface to the deeper horizon.

Spearman Rank Correlations (r_s) in Table 4 show that LAT significantly correlated with SOC_C (r_s = 0.15) implying higher SOC_C at higher latitudes. Surprisingly, MAP showed no significant correlation with any of the soil variables investigated in the current analysis but MAT correlated significantly and negatively with SOC_C (r_s = −0.14) and Fe (r_s = −0.35). Soil pH correlated significantly and positively with soil Clay, Chroma and Value ($0.18 \le r_s \le 0.23$) but negatively with SOC_C (r_s = −0.24). The Clay-Chroma correlation was significantly positive (r_s = 0.19); however, the negative correlation between Clay and SOC_C (r_s = −0.35) was rather surprising. It was also surprising that both Chroma and Value correlated significantly but negatively with SOC_C with respective r_s of −0.26 and −0.42. However, SOC_C was positively correlated to Fe (r_s = 0.21), while in turn Fe was also positively correlated to Al (r_s = 0.62).

These results also suggest that, unlike SOC_C, soil Clay and Al contents tended to decrease with the intensity of soil hydromorphy. Indeed, both soil clay (Clay, %) and aluminum (Al, %) content showed a general decrease with the hydromorphy index (HI) (Fig. 3A–B). In Fig. 3A, Clay decreased from 27 to 6% (i.e. by 60%) from HI range of 0–15 to 15–30, and this was followed by a further 59% decrease from 15 to 30 to 20–50. Interestingly, the 0–15 and 15–30 HI ranges showed much greater variability of Clay content than the 20–50 range. In the case of Al concentration (Fig. 3B), the decrease from 0 to 15 to 15–30 was by 50% (from 2.2 to 1.1%), while the decrease from 15 to 30 to 30–50 HI range was by only 36%. Variability of Al was greatest in the 15–30 range, and least in the 20–50 range.

3.3. Variation of HI as a function of climatic factors

HI correlated highly with SOC_C and it could, thus, be used as a proxy for SOC_C spatial prediction, the latter being potentially rendered easier by the use of ancillary data such as MAP or MAT because of their worldwide availability. Moreover, trends could inform on the mechanisms involved in the formation of soil hydromorphic features. In general, HI tended to decrease with increase in both MAP (Fig. 3C) and MAT (Fig. 3D). HI decreased from 25 to 16 from MAP range of 120–800 mm $year^{-1}$ to 800–1200 mm $year^{-1}$, and to 7 at MAP 1200–220 m yr^{-1}. HI varied greatly within the MAP ranges, but variability was greater at 120–800 and 1200–2200 mm $year^{-1}$ than at 800–1200 mm $year^{-1}$. In the case of MAT, HI varied greatly in all the three MAT categories but there was a trend for higher HI to occur at lower MAT values. HI decreased from 14.8 for 6–15C $year^{-1}$ to around 8.5 for 15–22 and 22–30 °C $year^{-1}$ MAT ranges.

3.4. Multivariate analysis

The two major PCs of the PCA (Fig. 4) explained 70.6% of the dataset variability, with PC1 and PC2 accounting for 47.2% and 23.4%, respectively. PC1 was closely associated with HI, MAP and soil depth; but HI showed negative coordinates while MAP and soil depth had positive coordinates on this axis. PC2 was closely associated with MAT and Clay with both of them showing negative coordinates along this axis. The results in this PCA suggest a general increase of SOC_C with increasing HI, and decreasing MAP and soil depth. In contrast, there was a tendency for soil clay content and MAT to have very little impact on both soil hydromorphy and SOC_C.

4. Discussion

4.1. Links between SOC_C and soil hydromorphy

The study results showed a general increase of SOC_C with HI (Fig. 2A and Fig. 4), which was consistent with results from many studies across the world (e.g. Chaplot et al., 2001; Jungkunst et al., 2008; Maljanen et al., 2010; Wang et al., 2011). This relationship could be explained by the impact of soil saturation on the stability of organic C. Saturation of soils by water creates reducing conditions where microbial activities are also suppressed due to very low free O_2 levels. Decomposition of soil organic matter (SOM) decreases to very low rates in the absence of microbial activity because lack of O_2 reduces the metabolic activities of the aerobic soil organisms responsible for decomposing the SOM. In their study of soil CO_2 effluxes, Freeman et al. (2004) explained that higher CO_2 effluxes from soil C pools under drier conditions were a result of greater decomposition rate. Jungkunst et al. (2008), who evaluated CO_2 effluxes from seasonally saturated soils due to variation of the groundwater table, confirmed this with their results indicating higher CO_2 emissions when the soils were relatively drier. The highest concentrations of dissolved CO_2 occurred in waterlogged soils (Jungkunst et al., 2008).

The accumulation of organic C in waterlogged soils induces typical changes in soil color (Schoeneberger et al., 1998; Vepraskas, 1999; Vepraskas and Caldwell, 2008; Vepraskas, 2001), especially lower Munsell Value and Chroma. The long periods of water saturation also reduce soil metallic elements, e.g. Fe and Al, to forms that exhibit dark to greyish colours (USDA-NRCS, 2006) which are also associated with low Value and Chroma (Munsell codes for soil color often used as indicators for soil hydromorphy). Transfer of Fe from soils is a mechanism for soil evolution controlled by water regime, which create reducing conditions. While hydromorphic environments favour dissolution of Fe out of soil systems (Fekiacova et al., 2013), the transfer of Fe may not be a definitive loss because the mottled horizons, especially those formed by water level fluctuations, tend to develop localized Fe accumulation structures (Wilson et al., 2013). Moreover, the Fe depleted structures may also result in geochemical degradation of ferruginous matrices (Beauvais, 1999). Guimaraes et al. (2013) also used similar processes to explain the occurrence of *Gleysol* profiles in the Brazilian Amazon floodplains. It is, however, still important to point that soil color is also influenced by transformation and accumulation of SOM and that SOM accumulation occurs in both biologically active soils with well humified OM and poorly active acidic environments with slow humification (Zaiets and Poch, 2016).

4.2. Links between soil hydromorphy and other soil properties and feedbacks to SOC_C

Soil matrix chroma is a property used for characterising hydrological regimes and boundaries of wetlands (Atkinson et al., 1993). While wetland environments are generally expected to have soils with low Value and Chroma (i.e. dark colours) (Vepraskas, 1999), the high Value and Chroma values observed in some of the wetland profiles used in the current analysis were of young *Gleysols* impacted on, for example, by river pulses. Considerable time is required for hydric soil indicators to develop (Atkinson et al., 1993; Campbell et al., 2002). The high soil Value and Chroma categories were associated with low SOC_C (Fig. 2A and C, respectively) pointing to possible enrichment of these soil profiles by high activity clay minerals low in SOC. Such soils when impacted on by river pulses tend to exhibit dark colours despite being generally freely-drained and oxidized environments (Vepraskas and Wilding, 1993).

Despite the positive impact of HI on SOC_C, the HI correlated negatively with soil properties such as clay (Fig. 3A) and chemical elements Al and Fe (Fig. 4). The decrease of soil clay content with HI indicates the existence of at least two pedogenetic processes causing loss of clays in the wetlands, (i) ferrolysis and (ii) physical weathering. Ferrolysis is a geochemical destruction process pertaining to the removal of Fe hydroxides during water movement along soil macro pores. The clay particles may be removed together with Fe hydroxides and be elluviated (through argilluviation) to lower horizons (Wilding and Rehage, 1985). The other process is still linked to ferrolysis in that clay particles are decomposed during ferrolysis in acidic and seasonally wet soils (Van Ranst and Coninck, 2002). This process is based on alternating reduction and oxidation of Fe. Mineralogical instability occurs in seasonally wet-dry conditions because the acids produced by decomposing organic matter, (under reducing conditions) generate soluble Fe^{2+} which moves through the soil and precipitates as $Fe(OH)_3$ (Espiau and Pedro, 1983). Aluminum is also dissolved out of soil systems in the same way because its solubility is highly dependent on pH (Bloom et al., 1979). The Al is mobilized into the aqueous form under acidic conditions, but precipitate to form amorphous Al hydroxides, Al-silicate secondary minerals, and/or Al–organic complexes under high pH conditions (Li and Johnson, 2016). Interactions between Al and SOM are directly linked to acidic functional groups on SOM surfaces (Bloom et al., 1979). The Al can be available for co-precipitation with silica to form Al silicates, which correspond with negative correlations between SOC_C and Al. In general, seasonally waterlogged soil pH decreases in the oxidized phase and rises in the reduced phase (Brinkman, 1969). The strongly positive correlation between soil clay content and pH ($r_s = 0.21$; Table 4) cements the role of ferrolysis, which becomes dominant under acidic conditions, in the destruction and transfer of clay particles. This also makes the positive correlations between soil clay and Al in the study results plausible because Al elevates soil pH through the presence of Al^{3+} and H^+ (Machado, 2014).

While our results point to negative correlations between SOC_C on one hand and Fe and Al on the other (Table 4), some study results are contrary to this (e.g. Wagai and Mayer, 2007; Kaiser and Guggenberger, 2000; Wilson et al., 2013; Akter et al., 2016). They explain that Fe hydroxides in water saturated soils play a role in stabilizing SOM through complex stabilization mechanisms by sorptive processes, or by the formation of organo-mineral complexes including SOM–Fe oxide–-clay associations (Kaiser and Guggenberger, 2000, Wagai and Mayer, 2007). Wilson et al. (2013) also pointed that strongly crystalline Fe hydroxide features are able to sequester OM in Fe–SOM complexes via encapsulation at the core of the Fe nodules. The anaerobic bacteria in waterlogged soils were also reported to increase the bond between Fe hydroxides and SOM (Akter et al., 2016). Similar results relating to Al are also available in literature; however, poorly crystalline Fe and Al hydroxides are more important than strongly crystalline mineral hydroxides for the stabilization of SOM (Spielvogel et al., 2008). This meta-analysis points to relatively low impact of organic C—Fe bonds on C accumulation in hydromorphic soils.

4.3. Factors controlling soil hydromorphy: towards better understanding of C stabilization in wetland soils

While soil hydromorphy would generally be anticipated to increase with precipitation due to amount of available water, our results show negative correlations between HI and climatic factors (Figs. 3 and 4). This may be explained in terms of weathering and its impact on drainage properties of soils. In general, high precipitation and temperature enhance weathering processes giving rise to well-structured soils with good drainage properties, which may explain their low hydromorphic intensity. In addition to climate, many other interactive controls of a geologic, geomorphologic and pedogenetic nature influence the development of aquic conditions and hydromorphy in soils. On the basis of topography, hydromorphic soils are normally found in valleys surrounded by hills with no permeable subsoil horizons to allow lateral flow and seepage downslope (Bouma, 1983). The aquic conditions may also be due to lithogenic discontinuity (Putzer, 1984), denudational events (Sombroek, 1990) and geochemical equilibrium between dissolution and precipitation during soil evolution (Dubroeucq and Volkoff, 1988). Geochemical landscape denudation over extensive geological periods in tectonically stable areas can cause soil surfaces to approach the top of, for example, a regional groundwater table leading to prolonged water saturation. Tectonic factors may also reduce soil drainage efficiency by promoting relative rises in base levels through tilting (Dubroeucq et al., 1999) and/or subsidence of large spatial extent.

Latitude is also an important factor controlling C stability in wetlands because landscape position and climate change control waterlogging conditions resulting in distinct SOM decomposition. For example, climate change induced thawing of permafrost in high latitudes is causing partially and/or seasonally waterlogged soils exposing previously frozen C to aerobic decomposition under more elevated temperatures, which increase CO_2 and CH_4 emissions by > 5 times in comparison to the same soil under anaerobic conditions (Gruber et al., 2004; Schuur et al., 2008). In the lowlands, the dissolved and particulate C exported by erosion from thawed soils can be deposited in aquatic environments (Walter et al., 2006).

Considerable fluctuations in soil spatial patterns due to differing drying rates following high rainfalls have also been reported (Guo et al., 2016). Soil drying is highly influenced by evaporation conditions (air temperature, humidity and wind speed) and its topographical location. Soil properties change due, for example, to changes in chemical element content, type and quantity of organic compounds (Vepraskas and Wilding, 1993; Peterschmitt et al., 1996; Montes et al., 2007) and may also alter the hydric systems. Such processes may involve lateral soil transformation where the hydromorphic properties develop in matrices of lateritic soils triggered by sharp structural, textural and mineralogical changes.

There are also anthropogenic factors which contribute to climate change and subsequently influence evapotranspiration potential and hydrology (Sombroek, 1990). Soil use may be considered a sensitive factor of color change associated with organic matter; for example, that land uses related with artificial hydromorphic soils such as rice cultivation keep high soil C levels due to the reducing character credited to long artificial flooding. Mineral soils of wetlands may also not have the organic horizon due to annual tillage operations (Ewing et al., 2012). These interferences contribute to the development of hydromorphic environments that result in increased water saturation and temporary immersion which reduce decomposition rates of organic matter in soil developed in topographical depressions.

5. Conclusions

In this study of 122 soil profiles from 29 studies performed under temperate, humid, sub-humid, tropical and sub-arctic conditions, our main objective was to evaluate the global variations in wetlands SOC content (SOC_C) and to relate this to key soil and environmental factors such as soil texture, intensity of soil hydromorphy, content in metals and climate. Three main conclusions can be drawn. The first one is that SOC_C whose average was 53.5 ± 15.8 g C kg^{-1} showed a range between 1 and 540 g C kg^{-1}. The second conclusion was that soil carbon content in wetland soils highly increase with increase in soil hydromorphy. The third one was that intensity of soil hydromorphy and thus SOC_C decreased with increasing mean annual precipitation. This was surprising as high precipitation would induce a greater occurrence of soil saturation by water but could be the result of greater oxygen input to the system or a greater soil drainage as rock weathering is enhanced. These results on the global variations in SOC_C in wetlands soils could inform land use planners and policy makers on the level of SOC_C to expect in a wetland of a given environment and thus the sequestration potential of degraded wetlands or the potential losses from natural wetlands when submitted to changes in land use or management. There were however several limitations to this study which were inherent to data availability. First of all, despite numerous studies on wetland soil carbon, there were very few studies simultaneously reporting on SOC_C and soil hydromorphy and more needs to be done to report on other important factors such as land use, geological bedrock and topographic conditions for a more holistic assessment of the controls of SOC_C level in wetland soils.

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